Evaluation of New GRACE Time-Variable Gravity Data over the Ocean

Don P. Chambers
Center for Space Research, The University of Texas at Austin
chambers@csr.utexas.edu

Abstract. Monthly GRACE gravity field models from the three science processing centers (CSR, GFZ, and JPL) are analyzed for the period from February 2003 to April 2005 over the ocean. The data are used to estimate maps of the mass component of sea level at smoothing radii of 500 km and 750 km. In addition to using new gravity field models, a new filter has been applied to estimate and remove systematic errors in the coefficients that cause erroneous patterns in the maps of equivalent water level. The filter is described and its effects are discussed. The GRACE maps have been evaluated using a residual analysis with maps of altimeter sea level from Jason-1 corrected for steric variations using the World Ocean Atlas 2001 monthly climatology. The mean uncertainty of GRACE maps determined from an average of data from all 3 processing centers is estimated to be less than 1.8 cm RMS at 750 km smoothing and 2.4 cm at 500 km smoothing, which is better than was found previously using the first generation GRACE gravity fields.

1. INTRODUCTION

In a recent study, I analyzed the initial release of monthly gravity field models from the GRACE mission over the ocean, after converting to equivalent sea level and smoothing over a radius of 1000 km [*Chambers*, 2006]. By comparing these monthly maps to ones computed from Jason-1 altimetry corrected for non-mass steric variations from the World Ocean Atlas 2001 (WOA01) database [*Stephens et al.*, 2001], I concluded that the mean accuracy of the smoothed GRACE maps was about 2.3 cm RMS, and between 3 and 4 cm RMS in the tropical Pacific. This accuracy appeared to be sufficient to recover significant ocean mass variations in higher latitudes, but it was unclear how well mass variations in the low- and mid-latitudes could be recovered with GRACE. Several problems in using the GRACE data were noted, including the lack of an ocean pole tide correction and very large errors in the measured $C_{2,0}$ gravity coefficient. Although I proposed methods to correct for these, I cautioned that the optimal solution would be for these to be fixed in the processing by the GRACE project, and not at the user level.

Since the publication of *Chambers* [2006], the two central GRACE Science Data System (SDS) centers (The Center for Space Research (CSR) and GeoForschungsZentrum (GFZ)) as well as the validation center (Jet Propulsion Laboratory (JPL)) have changed several background models and processing strategies in order to improve the gravity coefficients. In particular, an ocean pole tide model was incorporated in the processing and a new ocean tide model was used. In addition, other dynamical orbital parameters were estimated in order to improve the determination of the $C_{2,0}$ coefficient. In this article, I will evaluate these new monthly gravity field solutions in the same way as in *Chambers* [2006], by comparing with predicted ocean mass signals from steric-corrected Jason-1 altimetry.

It has also recently been demonstrated by *Swenson and Wahr* [2006] that the error characteristics of the GRACE gravity coefficients are not random for high degrees and orders, which has been assumed for the Gaussian smoother used to compute maps [*Wahr et al.*, 1998]. They found systematic errors in the higher order coefficients that tend to be different between odd and even degree coefficients for the same order. These errors propagate into north-south “stripes” when maps of equivalent water level or geoid are computed, depending on the level of smoothing used. In order to reduce the appearance of these “stripes” in maps, the GRACE gravity coefficients have had to be smoothed over relatively large radii (~1000 km or more). *Swenson and Wahr* [2006] have proposed a method to filter the GRACE coefficients in order to reduce these systematic errors, which they refer to as a “correlated-error filter”. Their results suggest a dramatic improvement in the ability of GRACE to resolve shorter wavelength features of mass variability when this type of filtering is applied. I will also include a similar filter in this analysis, and show what effect it has.

2. DATA PROCESSING

CSR, GFZ, and JPL all use different algorithms to compute gravity field coefficients from the raw GRACE observations, although they have agreed to use many similar background models. For the latest generation of gravity field models, the most significant differences have been the use of a new mean gravity field model determined using more GRACE data, the inclusion of an ocean pole tide as a background model, and the use of a new ocean tide model (FES2004) extended to higher resolution than previously. In addition, all centers modified their computation strategy in order to better estimate the $C_{2,0}$ coefficient. All tests indicate that GRACE now observes the $C_{2,0}$ coefficient as well as one determined from a satellite laser ranging (SLR) analysis [*J. Ries*, personal communication, 2006], so it is no longer necessary to substitute values from the SLR analysis as recommended in *Chambers* [2006]. Readers who are interested in the exact changes from the original release to the new release are advised to read the *Processing Standards* documents on the data archive site (ftp://podaac.jpl.nasa.gov/grace/doc).

The SDS centers have also designated the new releases slightly differently. CSR and JPL both refer to their newest models as Release-02 (RL02), while GFZ refers to theirs as Release-03 (RL03), as they had an interim release while CSR and JPL did not. However, CSR_RL02, JPL_RL02, and GFZ_RL03 should all be of comparable quality. One major difference is that CSR_RL02 fields use the ocean de-aliasing model used in the original release, while JPL_RL02 and GFZ_RL03 use a newer ocean model. This means that the CSR-RL02 coefficients will be biased relative to the JPL and GFZ solutions by the difference between the background models. However, after the monthly average of the appropriate model is added back to the coefficients in order to measure the full barotropic ocean variations [*Chambers*, 2006], there should be no systematic difference. In addition to adding back the de-aliasing model, a model for seasonal degree 1 (geocenter) variations is also used [*Chambers*, 2006].

New RL02 and RL03 gravity fields from February 2003 to June 2004 and November 2004 to April 2005 have been analyzed. Although GFZ_RL03 and JPL_RL02 solutions are available past April 2005, no CSR_RL02 solutions have been released past this point. There are no data for June 2003 and

January 2004 because of spacecraft problems during those periods. Although inter-satellite tracking data were taken during the 4 month gap between July to October 2004, the satellites were in a deep resonance and so gravity fields had significantly poorer accuracy and RL02-quality fields have not been released.

Before converting the GRACE gravity coefficients from the various centers into maps of equivalent sea level variations as described by *Chambers* [2006], the data have been filtered using the procedure similar to the one described in detail by *Swenson and Wahr* [2006] to reduce systematic errors in the higher order coefficients that tend to be different between odd and even degree coefficients for the same order. *Swenson and Wahr* [2006] found that the systematic errors began at approximately order 8 and that they are apparent at all higher orders. The basic idea behind the filter is to leave some *NxN* portion of the time-variable gravity field coefficients unchanged (where there are no obvious systematic errors) and fit a high-order polynomial as a function of degree for each order higher than N, with separate fits for odd and even degrees. This fit is assumed to be an estimate of the systematic error, and the filtered coefficient is then the original coefficient minus the fit. These filtered coefficients are then smoothed into maps of equivalent sea level using the equations given in *Chambers* [2006].

There are several degrees of freedom in the filter, from the value of N representing the unchanged portion of the time-varying gravity coefficients, to the value of N_{max} (the highest degree used in the fit) to the order of the polynomial used. *Swenson and Wahr* [2006] suggest one set of parameters based on their analysis and have used a running parametric fit. However, I wanted to optimize this for the ocean, and so analyzed several dozen permutations of the filter, altering the various components (N, N_{max}, order of polynomial, as well as the portion of the coefficients which are unchanged). In each case, I used the filtered CSR_RL02 coefficients to map sea level and computed global residuals with steric-corrected Jason-1 sea level maps as described in Section 3.0. The filter selected as optimal was the one with the lowest residual variance. The filter which has been implemented in this study keeps the lower 7x7 portion of the coefficients unchanged, then fits a 7th order polynomial to the remaining coefficients for each order (m) up to 50. The maximum degree (N_{max}) used in the fit is 80 for $m < 40$, while $N_{max} = m+40$ for m > 40. Only one polynomial is computed for each odd or even set for a given order, unlike the method of *Swenson and Wahr* [2006], which calculates multiple polynomials for each series. When converting to maps of smoothed water thickness anomalies, no filtered coefficients above degree/order 50 are used since the weighting functions with a radius of 500km or more are nearly zero for all degrees over 50. The effect of the filter on sea level maps from GRACE is dramatic. Figure 1 shows the maps for July 2003 from the CSR-RL02 coefficients at 500 km smoothing radius with and without the filter. Without the filter, the maps show unrealistic “stripes” that are an artifact of the systematic errors. With the filter, the maps appear much more realistic.

3. ANALYSIS OF RESULTS

After processing and filtering the GRACE coefficients from the three SDS centers as described in Section 2.0, the data were mapped to 21 monthly 1° grids with smoothing radii of 500 km and 750 km, masking ocean areas within the same radius of land to eliminate contamination from aliased hydrology variations [*Chambers*, 2006]. Maps of sea level from Jason-1 over the same period were also computed with the same smoothing radii as described in *Chambers* [2006]. The mean monthly steric variation was computed from the WOA01 database and smoothed, also as described in *Chambers* [2006]. The Jason-1 and WOA01 data are smoothed to the same level as GRACE data because one wants to compare mass anomalies averaged over approximately the same area. As the smoothing radius increases, the variance of the smoothed mass signal decreases. For instance, the global variance of the mass anomalies estimated from Jason and WOA01 data decreases from 7 cm^2 at 500 km smoothing to 5 cm^2 at 750 km and only 3 cm^2 at 1000 km. If one compared maps of mass variation from Jason and WOA01 at a smaller smoothing radius (say 300 km) to maps from GRACE at a larger radius (say 1000 km), a significant fraction of the residual variance would be due to shorter-wavelength variations that have been averaged out in the GRACE data, not necessarily error in the GRACE data.

As noted previously, I use WOA01 even though it is a mean climatology because it is still one of the few global databases with month-to-month variations, so observes the full range of seasonal variations [*Chambers*, 2006]. Many other databases of temperature, salinity, or steric sea level rely on averaging over considerable time-periods, from several months to several years, and so have too much temporal smoothing. They may also not contain data in many locations where there are sparse measurements, such as in most of the Southern Hemisphere. Because the WOA01 maps represent only the mean seasonal variation, the GRACE and Jason-1 maps were de-trended to reduce the influence of longer-period variations. This was done by estimating a bias, trend, and annual and semi-annual sinusoids for each 1° grid over the period from February 2003 to April 2005, then removing the bias and trend.

To estimate the uncertainty in the GRACE maps, they are compared with the Jason-1 and WOA01 maps for the appropriate month by computing the residual, Δ, where

$$\Delta(\phi,\lambda,t) = \begin{bmatrix} \Delta\hat{\eta}_{GRACE}(\phi,\lambda,t) - \\ \left(\Delta\hat{\eta}_{Jason}(\phi,\lambda,t) - \Delta\eta_{WOA}(\phi,\lambda,t)\right) \end{bmatrix}, \qquad (1)$$

η denotes the sea level anomaly (or equivalent water level for GRACE), ^ denotes a trend has been removed, t is the month, ϕ is the latitude, λ is the longitude. Smaller values of the residuals are interpreted to mean that GRACE agrees better with the expected signal represented by $\left(\Delta\hat{\eta}_{Jason}(\phi,\lambda,t) - \Delta\eta_{WOA}(\phi,\lambda,t)\right)$. The total variance over all t, ϕ, and λ is computed and analyzed. If one assumes no errors in the Jason-1 maps, no errors in the WOA01 maps, no non-seasonal steric variations, and that the difference represents the true ocean mass and barotropic variations, then the variance of the residuals represent uncertainty in the GRACE maps. In reality, this is only an upper bound, since both the Jason-1 and WOA01 maps have uncertainty and there are interannual steric variations that have not been accounted for. However, this type of analysis will show improvements in the GRACE data if they are changed but the Jason-1 and WOA01 data do not, assuming that the GRACE and Jason/WOA01 errors do not cancel.

Table 1 lists the variance computed from the residuals, where only the source or filtering of the GRACE coefficients is changed. The new processing standards and models for CSR_RL02 led to a significant improvement over CSR_RL01, even when the new correlated-error filter was not applied. The reduction in the variance is about 12% at 750 km smoothing and 16% at 500 km smoothing. Applying the correlated-error filter to the CSR_RL02 data in addition leads to a dramatic reduction in the variance at 500 km (more than 51%) with a more modest reduction at 750 km (17%). The statistics are similar no matter which SDS center provides the coefficients. Even more interesting is the fact that if the monthly maps from each center are averaged together to create a mean monthly map before computing the statistics, the variance decreases by another 8-9%. Averaging data from any two of the SDS centers has about the same effect. The lowest residual variance is found when all 3 centers' data are averaged, although the reduction is only another 5%. This suggests that averaging the monthly maps produced from each SDS center's data provides a better map of the ocean mass variability than using data from just one center. This is most likely due to reducing different random errors that arise in the gravity coefficients from each center using the same raw data but in different processing algorithms.

Figure 2 shows the local RMS statistics of the residuals over the 21 months based on this 3 center averaging. The mean RMS is 2.5 cm at 500 km smoothing and 2.0 cm at 750 km smoothing. The same assessment of CSR_RL01 data at 1000 km smoothing with no correlated-error filter was 2.3 cm RMS [*Chambers*, 2006], which means that the new GRACE gravity field solutions (with the correlated-error filter) have the same level of accuracy at 500 km as found before at 1000 km. This is even more compelling when one considers the global variance of the mass variations, which is 7 cm^2 at 500 km and only 3 cm^2 at 1000 km smoothing.

If a mean seasonal variation is computed based on sinusoid fits for each component of the residual (i.e., separate fits for GRACE, Jason-1, and WOA01), and then the residual is re-computed, the RMS statistics decrease significantly (Figure 3). The mean RMS of the seasonal fits is 1.4 cm at 500 km smoothing and 1.1 cm at 750 km smoothing. Part of the difference between Figure 2 and Figure 3 is undoubtedly interannual steric variations that are in the altimetry but are not removed with the WOA01 climatology or the estimated linear trend. The largest RMS values in Figure 2 tend to be in the tropical Indian and Pacific Oceans, as well as in the Agulhas current retroflection region and in the South Atlantic where eddies shed by the Agulhas propagate. It is known that there are significant interannual variations in all of these regions, related to ENSO in the Pacific [e.g., *Philander*, 1990], to the Indian Ocean Dipole [e.g., *Saji et al.*, 1999] and Monsoon [e.g., *Webster et al.*, 1999], and to interannual variations in the South Atlantic circulation [e.g., *Witter and Gordon*, 1999]. Thus, the high RMS values in these regions in Figure 2 are just as likely due to unmodeled interannual steric variations as errors in the GRACE data. The only region with unreasonably large residuals is near longitude 200°E in the South Pacific where ocean variations are small. This may be due to residual GRACE error in one or more of the months. Removing these regions from the analysis does decrease the variances listed in Table 1 slightly, by about 0.6 cm^2 at 500 km smoothing and by about 0.4 cm^2 at 750 km smoothing.

4. CONCLUSIONS

I have evaluated new, monthly time-variable gravity fields from CSR, GFZ, and JPL between February 2003 and April 2005 by comparing maps of smoothed mass density in terms of equivalent sea level with maps calculated from steric-corrected altimetry over the same time period. The statistics of the residuals represent an upper bound on the uncertainty of the GRACE data, as it ignores errors in both the Jason-1 and steric model and any non-seasonal steric variations that are in the altimetry but not in the steric-correction model.

The newest releases of gravity field solutions are approximately 12-15% more accurate than the original release, if unfiltered coefficients are used to create smoothed maps. However, if a new filter to reduce systematic errors is utilized [*Swenson and Wahr*, 2006], the error is reduced by a further 17% at 750 km smoothing and 51% at 500 km smoothing. Averaging results from the 3 processing centers (CSR, GFZ, JPL) reduces the variance of the residuals even further, by 11-13%. The estimated upper bound of uncertainty for GRACE ocean mass maps is 2.5 cm RMS at 500 km smoothing and 2.0 cm at 750 km smoothing for all frequencies when such an averaging is done, and 1.4 cm and 1.1 cm for seasonal fits at the same smoothing radii.

The type of analysis I have described has proven very useful for evaluating current GRACE gravity solutions. However, as the GRACE project continues to improve their understanding and processing of the data, it is likely that the accuracy of the GRACE data will improve to a level that statistics from this type of test will not decrease significantly. That is because one will more likely be quantifying the variance of interannual steric sea level variability that is unmodelled in the WOA01 climatology. To that end, future work will begin to use month-to-month steric-corrections based on measurements made from the new Argo array at approximately the same time as the GRACE and altimeter measurements. [*Gould et al.*, 2004].

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5. REFERENCES

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Table 1. Variance of residuals (Equation (1)) in cm^2 for 21 months in common. In all cases the Jason-1 and WOA01 maps were identical. The only maps that changed were those calculated from GRACE.

Source of GRACE Data	Smoothing Radius 500 km	Smoothing Radius 750 km
CSR_RL01, unfiltered[1]	18.4	6.5
CSR_RL02, unfiltered	15.4	5.7
CSR_RL02, filtered	7.5	4.7
GFZ_RL03, filtered	7.3	4.6
JPL_RL02, filtered	7.3	4.6
Average of CSR & GFZ	6.7	4.1
Average of CSR & JPL	6.7	4.2
Average of JPL & GFZ	6.8	4.2
Average of CSR, GFZ, JPL	6.5	4.0

[1] using corrections for ocean pole tide and $C_{2,0}$ as described in *Chambers* [2006].

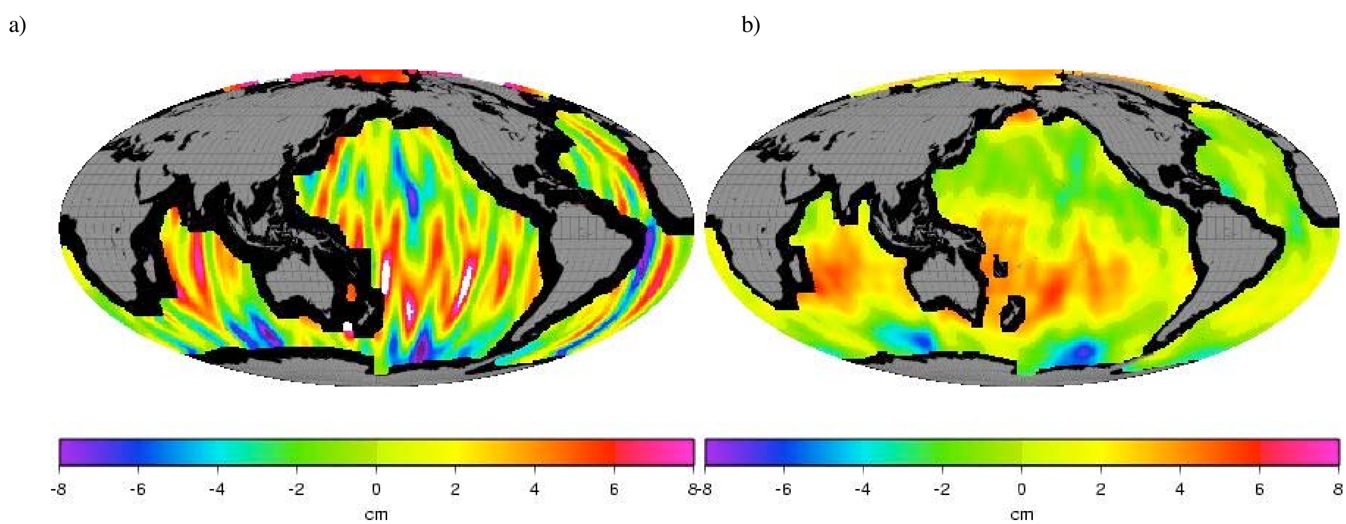

Figure 1. Sea level maps from the CSR_RL02 gravity coefficients for July 2003, smoothed over a radius of 500 km a) without using post-processing filter, and b) using filter.

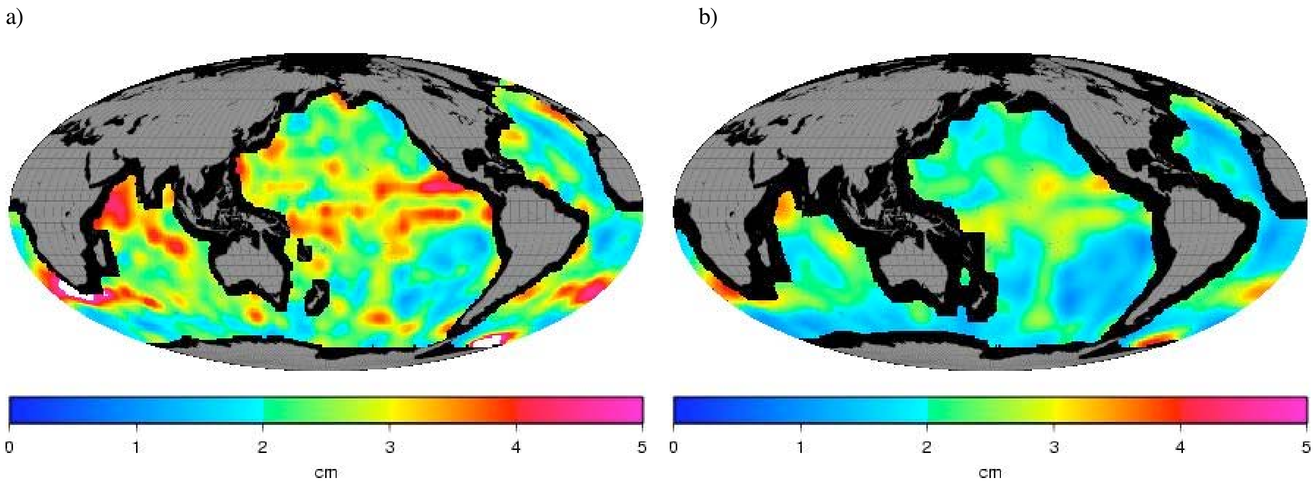

Figure 2. RMS of residuals for a) 500 km smoothing and b) 750 km smoothing.

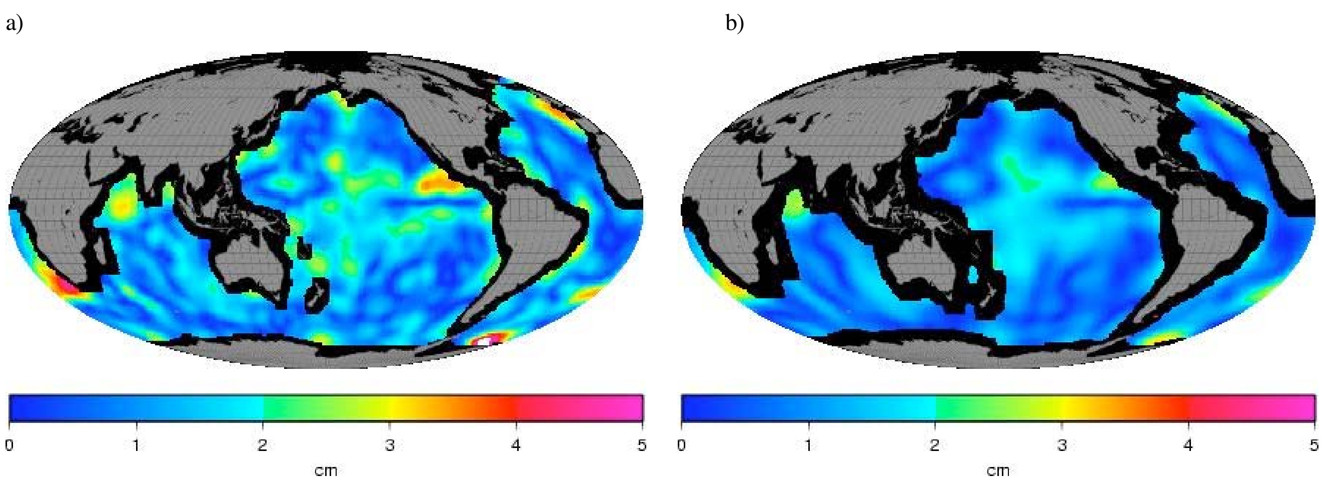

Figure 3. RMS of residuals computed from seasonal fits for a) 500 km smoothing and b) 750 km smoothing.